

# WHERE ON EARTH IS THE CRUST?

Mineralogical data and comparative planetology suggest a controversial hypothesis—that much of the Earth's crust may be buried within the mantle.

Don L. Anderson

Astronomers confidently speak of the chemistry of stars, galaxies and even interstellar space, but curiously, we are grossly ignorant of the composition of our own planet. "Whole Earth" scientists work under an extreme handicap compared with investigators in most other disciplines. Physicists, astronomers and biologists can formulate general theories and design experiments on the basis of many examples. By contrast, there is a single Earth, the result of a unique experiment, and we have direct access only to its outer portions. (Another obvious "singular" object of scientific study is the universe itself.)

Earth scientists long suspected that among the bodies in the solar system, our planet has unique geological characteristics; the era of planetary exploration, which has now established the discipline of comparative planetology, has amply confirmed this suspicion. Cataloging the similarities and differences among the planets gives us the raw material for a yet to be developed general theory for the origin and evolution of planets, including Earth. We know already that the patchwork of theories, laws and amendments that codifies our existing ideas about the Earth does not govern the detailed structures of other planets. Rather than further amend these "laws," it is now an appropriate time to start afresh—even if this means abandoning the beliefs and assumptions that most Earth scientists "know" to be true.

Geology is the dermatology of the outer skin of the Earth, a skin that we now know is extremely thin compared with those of our sister planets and that is not

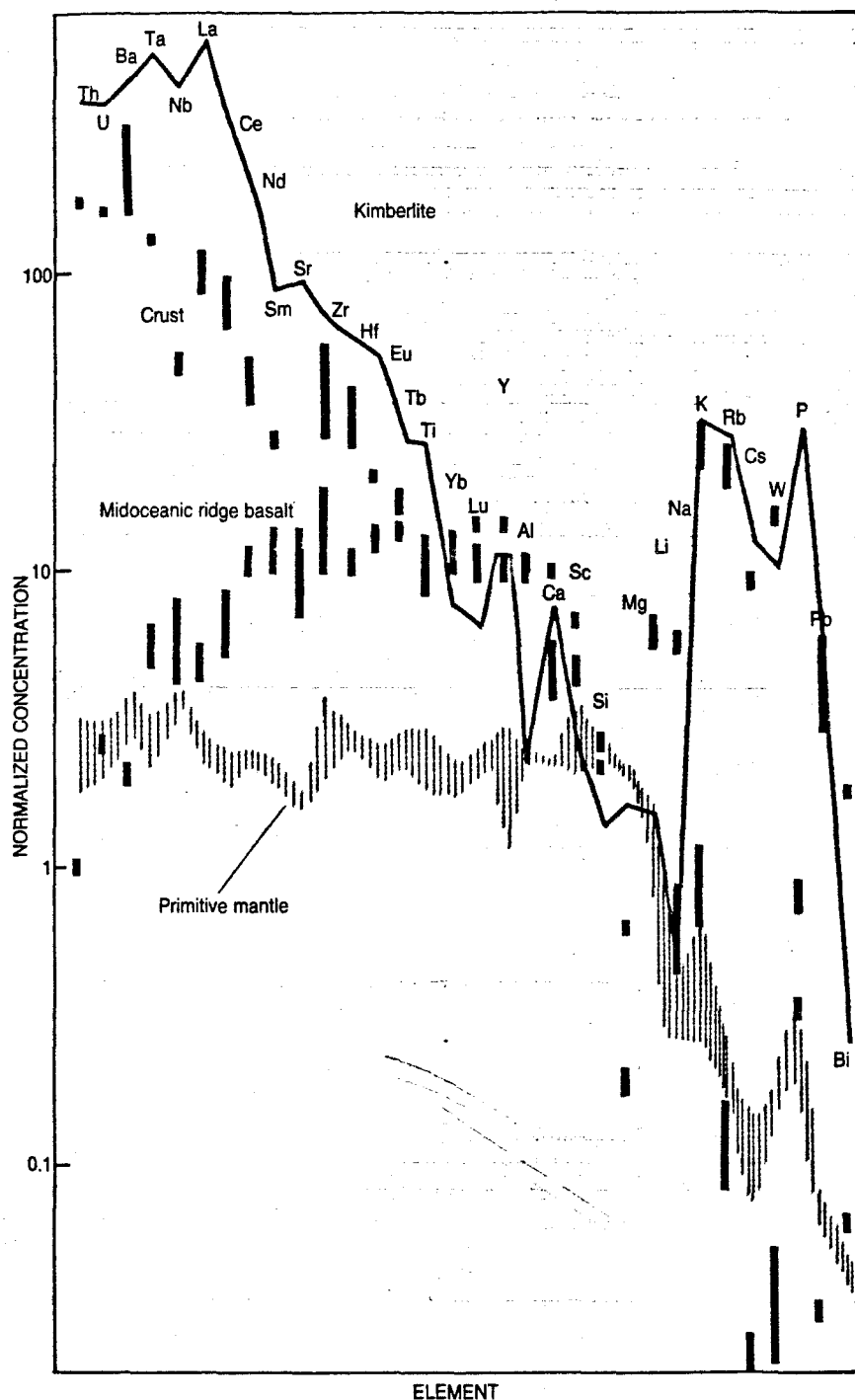
representative of the chemistry of the entire Earth. One of the oldest branches of geology is stratigraphy, the science of sedimentary layers near the Earth's surface. The distinctive properties of strata can be identified by eye or by seismic-wave detectors. Seismologists have now pushed this science to include the whole planet. The deeper one goes, the harder it is to detect strata, and so the tendency has been to think of the deep mantle of the Earth as a homogeneous mass having the same chemistry as the shallowest part of the mantle. Convective mixing is often invoked as the great homogenizer. The thinness of the crust and the stratigraphy of the mantle are two first-order observations bearing on the composition and evolution of the Earth.

## Composition of the Earth

Direct evidence for the composition of the Earth's interior is sparse. Earth's mass, radius and moment of inertia give us the average bulk density of the planet and a rough idea of the mean density of the core and mantle. Seismic determinations of the structure of the mantle and the radius of the core make it possible to refine these estimates, but this information is not adequate to allow us to determine the chemical composition without ambiguity. Indirect evidence or extraterrestrial arguments are needed.

Earth is presumed to have formed at the same time, and out of the same material, as the Sun, the other planets and meteorites. We know the composition of both the Sun and meteorites very well from spectroscopic analyses. But the Sun is made up primarily of volatile gases such as hydrogen, helium and nitrogen, while the Earth is made up primarily of rocky, dense, refractory material composed mainly of oxides of magnesium, silicon, iron,

**Don L. Anderson** is director of the Seismological Laboratory of the California Institute of Technology, in Pasadena. He is president of the American Geophysical Union for 1988-90.



**Bulk elemental compositions** of various components of the Earth as estimated from terrestrial and extraterrestrial data. The values have been normalized to abundances in carbonaceous chondritic meteorites. Earth is a well-differentiated body, as is evident from the variation in the concentrations of elements. Kimberlites and the continental crust have the highest enrichment. Midoceanic ridge basalts, while enriched overall, are depleted in the largest ions relative to the crust and relative to the smaller ions. The area labeled "primitive mantle" is the combination of crustal and mantle components that gives the best fit to ratios of refractory minerals in carbonaceous chondrites. The elements are arranged generally in order of decreasing abundance in the mantle-crust system, except geochemically similar elements are grouped together. The major rock-forming elements (Al, Ca, Si, Mg) and compatible elements (Sc, Li) are grouped together at center right, followed by the alkali metals, W and the volatile metals (P, Pb, Bi). Elements such as Th, U, Ba, Ta and the rare earth elements tend to be concentrated in melts; this tendency decreases as one progresses from La to Lu. **Figure 1**

aluminum and calcium. Clearly, the Sun and the giant planets collected most of the gaseous material in the solar system, while the smaller planets, near the Sun, collected mainly the rocky material. Because they most probably come from our solar system, meteorites, which bombard the Earth, provide us with low-cost samples from which to estimate the condensable content of solar system material. Meteorites are grouped into two general categories: stones and irons.

The average density and moment of inertia of the Earth can be nicely satisfied with a stony mantle and an iron core. If we make a few plausible assumptions we can regard the mantle as composed mainly of  $\text{SiO}_2$  and  $\text{MgO}$ , organized into more complex molecules, or minerals, such as enstatite, forsterite and their high-pressure forms, and

the core as composed mainly of iron. (Technical terms used in this article are defined in the glossary on page 45.) The next-most-abundant materials in the mantle are minerals that include  $\text{Al}_2\text{O}_3$ ,  $\text{CaO}$  and  $\text{FeO}$  in their formulas. Because the core's density is not quite right if we take it to be pure iron, we need a small amount, roughly 10%, of something lighter in the core, such as oxygen, sulfur, carbon, helium or hydrogen. The very refractory elements, such as aluminum, calcium, tungsten and titanium, and the platinum-group metals probably occur in the Earth in so-called cosmic ratios, because they are difficult to separate in solar system processes, although they do become separated by processes in planetary interiors such as melting and convection. From these considerations we can conclude that the Earth could

**Table 1. Estimates of solar system composition (relative to 1000 atoms of Si)**

	Solar composition <sup>1</sup>	"Cosmic" abundance <sup>2</sup>
Mg	1089	1075
Al	83.7	84.9
Si	1000	1000
Ca	82	61.1
Ti	4.9	2.4
Fe	1270	900

<sup>1</sup> From solar energetic particles, both coronal and photospheric. Data from ref. 4.

<sup>2</sup> Mainly from chondrites. Data from ref. 1.

have a Ca- and Al-rich crust at least several hundred kilometers thick and an Fe-rich core containing about one-third of the total mass of the Earth. Earth has the latter but not the former.

There is little dispute among Earth scientists regarding the aspects of the composition of the Earth's interior discussed so far. But other questions are extremely controversial:

- ▷ Is the mantle homogeneous in composition, or has it evolved into a series of distinct regions, with the lighter material generally concentrated in the crust and upper mantle and the denser material generally concentrated toward the interior, in the lower mantle and core?
- ▷ What fractions of the Earth are made up of the volatiles, the refractories and the in-between elements such as the major rock-forming oxides MgO and SiO<sub>2</sub>?
- ▷ Has the whole mantle, or just part of the mantle, been involved in the events leading to formation of the crust?
- ▷ Are the chondritic and solar abundances of the refractory elements the same?

It is generally assumed that the Sun, the planets and chondritic meteorites, having all derived from the solar nebula, contain approximately the same ratios of the refractory condensable elements. Thus elemental abundances in carbonaceous chondrites are usually taken as an appropriate guide to the composition of the condensable material in the solar system and the planets. Both chondritic and solar values are used in compilations of "cosmic," or solar system, abundances.<sup>1</sup> From these cosmic abundance tables one can generally develop satisfactory models of the Earth.<sup>2</sup> The inferred FeO

content of the mantle for a "chondritic" Earth, however, is very low, because nearly all the iron resides in the core. Seismic data, in contrast, suggest a value near 14% for the FeO content of the lower mantle,<sup>3</sup> and other silicate bodies also have high FeO contents.

Recently the compositions of the Sun's corona and photosphere have been redetermined.<sup>4</sup> The Fe atomic abundance by number is about 40% higher than in previous measurements, and the abundances of Fe, Ca and Ti relative to Si, Al and Mg are 30–40% higher than the chondritic values (see table 1). The new values suggest that carbonaceous chondrites may not be representative samples of the solar nebula and that the terrestrial planets, which are assumed to have roughly solar abundances, may be more Fe, Ca and Ti rich than previously supposed. With these new values the Earth's mantle would have more FeO and more CaMgSi<sub>2</sub>O<sub>6</sub>, the diopside component of clinopyroxene.

If one assumes that the Earth accreted from a nebula of solar composition and takes the core composition to be roughly Fe<sub>2</sub>O, one can estimate the composition of the terrestrial mantle from these new solar abundances. That estimate for the core composition is close to the inferred eutectic, or minimum melting, composition of the Fe–O system at high pressure<sup>5</sup> and also has about the right density to satisfy the geophysical data for the core.<sup>6</sup> The core<sup>7</sup> contains 32.5% of the Earth's mass, and it is assumed that all the Fe remaining after one accounts for the Fe in the core is in the mantle as FeO in silicates—along with fully oxidized Mg, Si, Al and Ca. The resulting mantle composition is given in table 2. The inferred FeO content of the mantle is about 15.7% by mass. This is much greater than the 8–12% FeO composition of upper-mantle lherzolites and basalts. The implication is that the lower mantle, which represents 70% of the mantle, is enriched in FeO compared with the upper mantle. The high inferred FeO content of the mantle is similar to estimates for the lunar and Martian mantles (see table 3) and about the same as the FeO content of the silicate portions of chondritic meteorites. It is also close to estimates from seismological data. Many authors have tried to rationalize the differences in FeO content between the Earth and these other bodies. It now appears that these exercises were unnecessary, and that the Martian, lunar and terrestrial mantles may be similar.

Once one has these estimates of the composition of an average mantle, the next step is to determine how the elements might have been rearranged by processes of planetary accretion and differentiation.

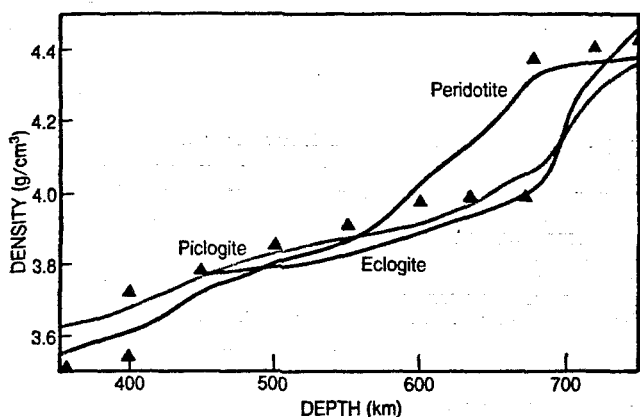
## Differentiation of the Earth

Clearly the Earth has a crust, a mantle and a core, so to first order it is a chemically differentiated and stratified body. If we liken it to a blast furnace, the crust is the lighter slag and the core is the iron at the bottom of the kiln. The Earth's crust averages about 20 km in thickness and reaches about 60 km under some of the higher mountain ranges. That continents are above sea level, that mountains stand tall and that the oceans are deep is due primarily to variations in crustal thickness: The crust is thin under oceans and thickest under mountains. The oceanic crust differs in chemistry from the continental crust and is denser. The continental crust is buoyant and acts like an iceberg. Compared with the rest of the Earth the crust is enriched in calcium, aluminum, sodium, potassium and a variety of other elements that are easily

**Table 2. Chondritic and solar estimates of terrestrial composition vs upper-mantle composition (percent by mass)**

	Chondritic	Solar	Lherzolite
<b>Mantle</b>			
MgO	37.7	32.7	42.2
SiO <sub>2</sub>	52.5	45.0	44.2
Al <sub>2</sub> O <sub>3</sub>	3.8	3.2	2.1
CaO	3.0	3.4	1.9
FeO	3.3	15.7	8.3
MgO/SiO <sub>2</sub> ratio	0.718	0.727	0.955
<b>Core</b>			
Fe <sub>2</sub> O	32.2	32.0	

Data from table 1 and from S. Maaloe, K. Aoki, *Contrib. Mineral. Petrol.* **63**, 161 (1977).



**Density vs depth** for several mantle constituents along a 1400 °C adiabat. The curve for eclogite approximately merges with that of piclogite at depths shallower than about 450 km. Triangles show the density variation of the Preliminary Reference Earth Model with depth in the mantle.<sup>7</sup> Note that the three rock types are all less dense than the top of the lower mantle and therefore will not sink into the lower mantle unless they are very cold. Because the major rock types of the mantle differ in density and melting temperature, they probably separate and form a layered system. A high FeO content for the lower mantle is one way to explain its high density. **Figure 2**

melted out of a rock and rise readily to the surface (see figure 1). The crust can be considered the "scum" that forms when a planet cools and freezes. There is enough crust-forming material—low-density, low-melting-point rock—in the Earth to form a crust ten times thicker than what is actually there. Considering the high energy and the melting associated with planetary accretion, we can expect the mantle also to be chemically stratified, unless the densities of the various products of melting and differentiation are so similar that they cannot separate.

One would think that seismology could easily resolve the possibilities. Seismologists have in fact discovered several first-order discontinuities in the mantle, and these may be chemical discontinuities just as the crust-mantle and mantle-core boundaries are. (Table 4 lists the major subdivisions of the Earth.) A problem is that such discontinuities can be due to phase changes as well.<sup>8</sup> To compound the problem, there are few laboratory data on seismic velocities in the high-pressure minerals that occur at great depths in the mantle, or even on the temperatures and pressures at which the phase changes occur. What data exist suggest that at depths greater than 400 km, the new phases (or high-pressure minerals) that are formed almost, but not quite, explain the seismic properties of the deep mantle. Whether the fit is good enough, given the uncertainties, is a contentious issue in geophysics. So-called *ab initio* calculations of elastic properties<sup>9</sup> favor an FeO-rich lower mantle, but they are an order of magnitude less precise than the seismic data.

The seismic data for the lower mantle are consistent with a chondritic, or solar, MgO to SiO<sub>2</sub> ratio and a high FeO content.<sup>3</sup> These are different from the values for the crust and upper mantle. The lower mantle is important in any mass balance calculation because it contains a large fraction of the Earth's mass (see table 4).

The high FeO composition of the lower mantle that we got earlier using the solar abundance data would effectively prevent mixing between the upper and lower mantles. This amount of FeO raises the density to a value higher than can be overcome by cooling of upper-mantle material. It also suggests that a subducting slab of crustal and upper-mantle material might hit bottom at about 650 km, the boundary between the upper and lower mantles and near the maximum depth for earthquakes. In the standard model of mantle dynamics, the mantle is homogeneous and cold slabs can readily sink into the lower mantle.

Another line of attack on the mantle composition problem uses actual samples from the mantle, of which we have several kinds (see figure 1). When part of the mantle melts, it forms a low-density magma that rises to the Earth's surface, mostly at midocean ridges, to form new crust. This magma contains information about the

composition of the mantle, but what fraction of the mantle it represents or from what depth it comes are both unknown. It is unlikely that the mantle melts very extensively to form basalt because the light magma that rises to the surface probably leaves behind a dense, unmelted fraction that represents most of the rock (at least this is the case in the current and generally accepted view, which attributes all basalts to less than 20% melting of the mantle).

At great depths, probably deeper than 200 km, the densities of melts and residual crystals are similar, so the propensity for melts to drain upward is greatly reduced. The density of a partially molten rock at such depths, however, is less than the rock's density before it melted. This is because the densest material in the rock, garnet, is eliminated upon partial melting. Therefore any part of the mantle that starts to melt becomes buoyant and rises toward the surface, becoming part of a gigantic convection cell. The decrease in pressure as it rises causes more melting and more buoyancy, so that the deeper mantle can lift itself "by its own bootstraps" into the shallow mantle. There the melt eventually separates as described above to form new crust. In such a dynamic situation we can no longer constrain the amount of melt to be small and cannot estimate the amount of solid material left behind. It could even be that a deep basaltic reservoir must be almost completely molten before it is buoyant enough to rise to the surface through the low-density, olivine-rich shallow mantle. In any case basalts are at best samples of only part of the mantle, and we are not even sure which part.

Komatiites are dense, high-temperature, high-MgO magmas characteristic of ancient (Archaean) terrains. Judging from their MgO content, crystallization temperatures and age, komatiites are the most "primitive" of magmas and should be more similar in chemical makeup to the source material than are other magmas. Table 3 shows a remarkable similarity between the composition of the oldest komatiites, from Barberton, South Africa, and our estimates for the primitive mantle. The similarities include high FeO contents and high CaO to Al<sub>2</sub>O<sub>3</sub> ratios. (The Barberton lavas are  $3.5 \times 10^8$  years old; younger rocks have quite different compositions.)

## Mantle fragments and exposed mantle

In addition to magmas, geologists have other samples of the mantle to work with. Some magmas contain fragments of solid material that were ripped out of the upper mantle during eruption. The presence of these fragments indicates that separation of solid rock and liquid magma is not always complete. Some fragments contain diamond crystals, demonstrating that they and the magma come

from deep within the upper mantle rather than from the crust. Most of the crystals, however, are the refractory minerals olivine and orthopyroxene.

The refractory solid fragments, although providing important clues about the mantle through which the magmas pass, have almost negligible mass compared with the amount of material that emerges from the mantle at midocean ridges, volcanic islands at oceanic "hot spots," and other volcanic centers. And since the fragments come mainly from depths less than 200 km, they do not tell us about the deeper mantle.

The most massive exposures of refractory upper-mantle material occur in some deeply eroded mountain belts. Olivine is the dominant material in these exposures, usually in a heavily altered form known as serpentinite. This observation has led geologists to the view that olivine, in roughly the composition  $(\text{Mg}_{0.9}\text{Fe}_{0.1})_2\text{SiO}_4$ , is the dominant component of the mantle. That olivine is a common mineral in meteorites has been taken as supporting evidence for this view, although meteoritic olivine is generally much more iron rich than the terrestrial variety. Olivine, particularly serpentinite, is lighter than the average mantle rock density and "floats" at the top of the mantle. It has a very high melting temperature and is left behind when basaltic magma finally separates from the upper mantle to form volcanoes. Much of the olivine in the upper mantle is probably left over from melting that took place early in Earth's history and is unrelated to the basalts that are currently erupting through it to the surface.

Thus we have in the Earth system a natural mechanism for forming two layers. The crust, mainly basaltic and formed by billions of years of melting and volcanic activity, is the lightest material in the system and forms the surface layer. Because of its intermediate density, olivine, the highest-melting-point, or most refractory, common mineral, is sandwiched between the crust and the remaining mantle, along with some orthopyroxene. Magmas from deeper layers pass through the olivine-rich layer on their way to the surface. Peridotites and lherzolites are common olivine-rich shallow-mantle rocks.

Basalt has an interesting characteristic. At low pressures it has a low density when molten, and it therefore rises to the surface. Cold basalt, however, converts to a dense rock called eclogite when it is deeper

than about 50 km in the mantle. Basalt attempting to rise beneath a thick cold plate, for example, a continent or an old oceanic plate, converts to eclogite and, being denser than the normal mantle and olivine, eventually sinks back into the deeper mantle. Oceanic crust shoved back into the mantle by plate-tectonic processes will suffer the same fate if it does not melt first. If underplating occurs, that is, if all the melt in the mantle does not make its way to the midoceanic ridges, the oceanic plate may be a basalt-olivine-basalt sandwich; the lower part of the sandwich converts to eclogite as the plate cools or sinks at island arcs, such as Japan, the Aleutians and Tonga-Fiji, and beneath such volcanic lines as the Cascades, in the US Northwest, and the Andes. This gravitational sinking of eclogite through the olivine-rich layer and normal mantle gives a mechanism for forming a third layer in the Earth system. It also gives a mechanism for offsetting the buoyancy of the crustal and olivine-rich layers.

The basic question is whether light basalt and dense eclogite can segregate from residual mantle or whether they are mixed back in. Most calculations to date treat mantle heterogeneities as passive, well-distributed, isodensity, isoviscous blobs of rather small dimension. Given these assumptions, the heterogeneities rather obviously eventually get stirred back in. Inhomogeneities of large dimension and density contrast tend to settle out. In a chemically stratified mantle it is difficult to entrain and homogenize material having quite different densities and viscosities.

The depth to which cold eclogite will sink—the "floor" of the system—is unknown but may be the same as the maximum depth of earthquakes, about 670 km. This is also about the depth of the boundary between the upper and lower mantles. If the lower mantle differs enough in chemistry from the sinking plate, then material will pile up or spread out at the boundary. ("Different enough" means that the intrinsic density of the lower mantle is higher than that of the cold slab material.) Eventually the subducted material will warm up and return to the ridge to complete the cycle. The buoyancy provided by melting and the negative buoyancy associated with the basalt-eclogite phase change affect convection in the mantle and may be the dominant driving forces in plate tectonics. This is an effect that has generally been ignored in modeling. Even if the upper and lower mantles remain

Table 3. Solar estimate of mantle composition vs compositions of mantle lherzolite, komatiites and pyrolite, and of lunar and Martian mantles

	Solar <sup>1</sup>	Lherzolite <sup>2</sup>	Komatiites <sup>3</sup>		Pyrolite <sup>4</sup>	Bulk Moon <sup>5</sup>	Mars <sup>6</sup>	Mars <sup>7</sup>
SiO <sub>2</sub>	45.0	44.2	44.9	44.8	45.1	43.4	44.4	41.6
Al <sub>2</sub> O <sub>3</sub>	3.2	2.0	3.1	3.6	4.6	6.0	3.02	6.4
FeO	15.7	8.3	13.5	12.0	7.8	13.0	17.9	15.9
MgO	32.7	42.3	33.0	34.9	38.1	32.0	30.2	29.8
CaO	3.4	2.1	3.8	3.7	3.1	4.5	2.45	2.4
CaO/Al <sub>2</sub> O <sub>3</sub> ratio	1.08	1.07	1.21	1.02	0.67	0.75	0.81	0.38
MgO/(MgO + FeO) ratio	0.68	0.84	0.71	0.74	0.83	0.71	0.63	0.65

<sup>1</sup> From solar energetic particles, both coronal and photospheric. Data from ref. 4.

<sup>2</sup> Average of 301 continental spinel lherzolites. Data from S. Maaloe, K. Aoki, *Contrib. Mineral. Petrol.* **63**, 161 (1977).

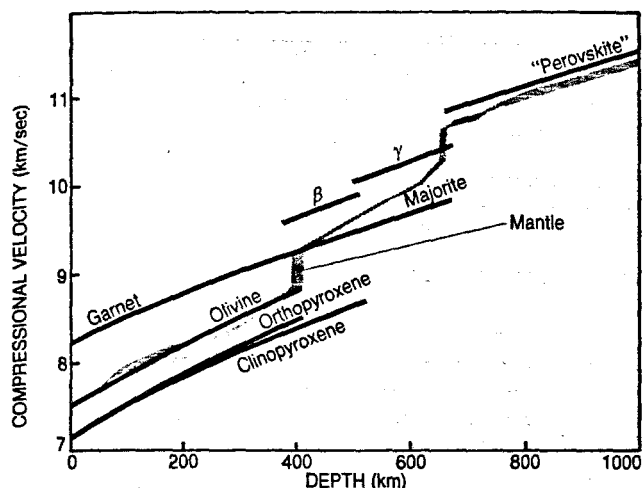
<sup>3</sup> Two samples from Barberton, South Africa. Data from N. T. Arndt, E. G. Nisbet, eds., *Komatiites*, George Allen and Unwin, Boston (1982).

<sup>4</sup> Data from A. E. Ringwood, *Composition and Petrology of the Earth's Interior*, McGraw-Hill, New York (1975).

<sup>5</sup> Data from S. R. Taylor, *Planetary Science: A Lunar Perspective*, Lunar and Planetary Institute, Houston, Tex. (1982).

<sup>6</sup> Data from G. Dreibus, H. Wänke, *Meteoritics* **20**, 367 (1985).

<sup>7</sup> Data from J. Morgan, E. Anders, *Geochim. Cosmochim. Acta* **43**, 1601 (1979).



**Compressional velocity vs depth** for major mantle minerals and their high-pressure forms (adapted from T. Duffy, D. L. Anderson, *J. Geophys. Res.*, in press). The high velocities in the high-pressure forms of olivine ( $\beta$  and  $\gamma$ ) show that the transition region must contain abundant clinopyroxene and may contain large amounts of garnet. These are the eclogite minerals, contained in high-pressure "crust." The lower mantle has a velocity slightly less than that of the orthopyroxene in the perovskite structure. Results from seismic models are indicated. **Figure 3**

separate, the lower mantle may influence the upper mantle by shear or, more likely, thermal coupling.

Figure 2 shows the densities at various depths of eclogite and peridotite, two possible components of the upper mantle and of subducting slabs. Note that their densities cross near 500 km and that both materials are less dense than the lower mantle. A cold eclogite slab can sink to about 600 km in a warm peridotite mantle, and a mixture of peridotite and eclogite can sink deeper still. This suggests that slab material accumulates in the upper mantle. Eclogite, or piclogite, being denser than peridotite over most of the upper mantle, particularly when cold, probably accumulates in the lower part of the upper mantle, the transition region.

The seismic velocities in the mantle help constrain the mineralogy and chemistry. Because the temperature and pressure derivatives of the elastic properties of the high-pressure phases are uncertain, it is difficult to calculate the velocities for the conditions that prevail *in situ* in the mantle. Results from one attempt are shown in figure 3. Olivine, orthopyroxene, clinopyroxene and garnet are the most abundant minerals in the upper mantle above 400 km. Near 400 km olivine converts to  $\beta$ -spinel and orthopyroxene converts to majorite, a garnet-like cubic mineral. From the figure it is obvious that these phase changes cause a much larger jump in velocity than actually occurs at 400 km and that the seismic gradient deeper than 400 km is much steeper than self-compression of homogeneous minerals. This is indicative of a smeared-out phase change, one that does not occur all at once at a given pressure. If these calculations are approximately correct, they indicate that there must be abundant clinopyroxene below 400 km to keep the velocities low and that there cannot be as much olivine and orthopyroxene in the transition region as in the shallow mantle. Garnet also can be present, since it has velocities similar to those

observed near 400 km. Modeling suggests an olivine-poor (somewhat less than 50%) and garnet-clinopyroxene-rich transition region—in other words, a basalt-eclogite-rich region, with abundant CaO and  $\text{Al}_2\text{O}_3$ . This region is floating on the dense, FeO-rich lower mantle.

## Comparative planetology

Now let us see how other objects in our solar system might help us to understand the Earth. The development of the theory of plate tectonics and the era of planetary exploration both started about 20 years ago. The plate-tectonic "revolution," which appeared to give a grand unifying hypothesis for the way a planet works, was soon found to work only for the Earth.

Venus, Earth's virtual twin in size and mean density, is dominated by a few continent-sized volcanic constructs, but otherwise has bland topography, with no evidence of globe-encircling ridge or trench systems indicative of plate tectonics. The Moon had a violent early history, complete with a global magma ocean and floating "rockbergs," but quickly passed into an old age characterized by a quiet, unchanging exterior and barely detectable signs of internal creaking and groaning. The whole Moon was occasionally reoriented on its spin axis by gigantic impacts. The Martian landscape is now dominated by a few, apparently ancient, massive volcanoes, which apparently also re-oriented the whole planet so that they now lie nicely astride the equator. The crusts of both Mars and the Moon appear to represent orders of magnitude more material, relative to the size of each body, than the Earth's crust, and on neither body is there any evidence of large-scale motions of surface plates. Superficially Mercury has a surface appearance like the Moon's, but it is by far the densest planet, having a large iron-rich core. Mars exhibits tensional stretch marks, while Mercury is dominated by compressional wrinkles.

What have we learned from our unique neighbors that might be useful in understanding the Earth? From the Moon we have learned that extensive melting, a magma ocean and a thick crust are likely early stages for a planet that is now solid. From Mars and the Moon we have learned that planets can tumble relative to their spin axes and thereby reorient their tidal bulges, causing the surfaces to rip and wrinkle. Mars is much smaller than the Earth and has a much lower density. When one corrects for pressure, Mars is either deficient in iron or contains more oxidized iron (that is, a higher FeO to Fe ratio) than the Earth. Mercury has much more iron, and the Moon much less, than the other planets. Mars, the Moon and many meteorites appear to have high FeO contents in their silicate portions, similar to the FeO content inferred for the lower mantle and much higher than that of upper-mantle rocks.

From Venus we have learned that a thick atmosphere and greenhouse heating can heat the surface to the point where it remains buoyant, preventing the foundering and sinking of the crust. Venus is slightly smaller than the Earth and has an extremely high surface temperature. These conditions combine to depress solid-solid phase changes and elevate melting transitions. As a result Venus, given its slightly lower density, may have the same composition as the Earth.

## A new Earth

Current theories for the origin and evolution of the Earth retain many elements that are holdovers from pre-plate-tectonic, pre-planetary-exploration concepts. The Earth is

usually assumed to have been assembled from a homogeneous mixture of cosmic material. Sometime after Earth formed, the story goes, the dense iron core settled to the interior and the crust was "sweated out" of part of the upper mantle. Most of the mantle is presumed to still be homogeneous, *primitive*, unmelted original material that is only now melting and providing basalts to the surface. This scenario implies either a cold origin or a stage of such vigorous convection and rapid cooling that the Earth's interior was homogeneous throughout. Both of these alternatives now seem exceedingly unlikely.

The extremely thin terrestrial crust might be used as evidence that most of the mantle has not been exposed to melting and therefore still retains the potential crust-forming material. However, the crust, thin as it is, contains such high proportions of elements such as potassium, rubidium, barium, uranium and thorium that most of these elements must have been removed from the mantle. Some magmas, notably kimberlite, also are enriched in these elements. This enrichment further reduces the amounts of these elements that one needs to localize in the remaining mantle material. This evidence points toward an extremely well-differentiated Earth, with a strong upward enrichment of the elements that are concentrated into magmas.

The oceanic lithosphere is about 8 km of basaltic material and tens of kilometers of buoyant, olivine-rich mantle material, both formed at midocean ridges. As the oceanic lithosphere moves away from a ridge, it cools and becomes more dense. Basaltic material added to the base of the lithosphere as it ages eventually converts to eclogite and contributes to the negative buoyancy of old oceanic plates. An unknown part of the lower oceanic lithosphere is basaltic in composition but it is unlikely that the crust is the only basaltic part of the oceanic lithosphere.

The continental crust averages about 30 km in thickness and is  $\text{SiO}_2$  rich and MgO and FeO poor compared with the oceanic crust; it is therefore much less dense. The thickness of the continental plate is unknown, but stable shield areas have about 120 km of olivine-rich and garnet-poor material, which makes the plate buoyant relative to normal mantle, even when cold. Continental lithosphere, at least in stable shield areas, appears to be immune to subduction, and the oldest rocks are found in these areas. The thickness of the continental crust was probably established early in Earth's history and may be related to the depth of the eclogite phase change shortly after the freezing of the magma ocean.

Why is the Earth's crust so thin when the potential crust in the mantle would easily make a crustal layer more than 200 km thick? Both the Moon and Mars, very small objects by terrestrial standards, have crusts of roughly the expected size. The answer is that a thick crust is impossible on an Earth-sized planet with a cold surface because of the aforementioned basalt-eclogite phase change. The base of any crust thicker than about 50 km,

approximately the maximum thickness of crust on the Earth, converts to dense eclogite and behaves not as crust but as particularly dense mantle. Not only is it dense, and therefore not buoyant, but it has mantle-like seismic properties. (Crust is identified by its low seismic velocities.)

If surface temperatures on the Earth were as high as they are on Venus, the lithosphere would probably never get cold enough to subduct. In addition, the eclogite phase change would occur at greater depth and a thick basalt crust would melt before it converted to the dense phase. Basalt pushed into hot mantle would return to the surface instead of sinking through the upper mantle. This may have been the situation in early Earth history, when the mantle was much hotter than it is at present. Venus may provide a good analog to the early, hot Earth. On the Moon a deep magma ocean ultimately formed a thick basalt crust lying over a bed of refractory olivine crystals—basalt rises; olivine crystals sink. (This would also happen on a small planet.) The Moon is so small that the basalt-eclogite transition never took place.

On an Earth-sized planet with a cooling, crystallizing magma ocean, a thin basalt layer forms at the surface, a thick eclogitic layer forms at depth and an olivine-rich layer forms at intermediate depth. Most of the "crust" is buried. A thick, hot crust slowly converts to dense eclogite as it cools, so that eventually it loses its buoyancy and sinks. Such a first-order catastrophe would certainly erase the early geological record and explain the missing  $0.5 \times 10^9$  years between Earth's formation and the oldest known rock. Perhaps there were several such episodes of "crustal cannibalism" before the upper mantle cooled sufficiently for thick continental rockbergs to become stable. Oceanic crust is still being constantly digested and regurgitated. The source and sink of oceanic lithosphere may be deep in the mantle, in the transition region.

The Earth, of course, supports plate tectonics, a fact we were learning when we found that other planets did not. We do not yet understand why plate tectonics exists on the Earth, or what drives the plates. Simple convection theory identifies the plates with the thermal boundary layers of mantle convection cells that are dragged down by the negative buoyancy of descending plates at deep-sea trenches. This ignores the chemical buoyancy of the crust and uppermost mantle and the high strength and viscosity of the plate. There exists, at present, no satisfactory theory for the initiation of subduction. The plate-tectonic theory is a descriptive rather than a physical theory. Any physical theory must explain why the Earth behaves one way and the other planets other ways. The most obvious differences among Earth, Mars and the Moon are in size. A small planet is likely to cool quickly and have a thick, cold outer shell that will shut off plate tectonics. A large planet, such as the Earth, not only starts out hotter but cools more slowly, particularly if it is chemically stratified. In a large planet pressure-induced phase changes may trap material falling through its mantle. A small planet is more likely to have a chemically homogeneous mantle. A hot planet will have a convecting interior—a prerequisite for plate tectonics. But even though Venus, a near twin of the Earth in size, is so hot at the surface that lead would melt there, it appears not to have plate tectonics. Apparently, a planet can be too hot to support plate tectonics.

Most of Venus has gentle, rolling plateaus with little topographic relief, but there are a few Australian-sized "continents" that rise more than 10 km above the plains. In contrast to the Earth, there is an excellent correlation of elevation and gravity. This shows that there is low-density material extending to great depth under the

**Table 4. Subdivisions of the Earth (fraction by mass)**

Crust	0.005
Upper mantle	0.103
Transition region	0.075
Lower mantle	0.492
Outer core	0.308
Inner core	0.017

Data from ref. 7.



## Glossary

**Basalts**—Melts or rocks crystallized from melts. Most magmas are basaltic in composition. Basalts represent the low-melting fraction of the mantle and are rich in Si, Ca, Al, Ti, Na and K. They have low density at upper-mantle pressures and tend to rise to the surface to form crust, or to shallow levels in the mantle if eruption is prevented. At high pressure basaltic melts and eclogite are denser than residual mantle rocks such as lherzolite.

**$\beta$ -Spinel**—A high-pressure form of olivine, stable between about 400 and 500 km depth in the mantle;  $\gamma$ -spinel is the stable form below about 500 km.

**Chondritic material**—Material having some of the chemical properties of the commonest meteorite class, chondritic meteorites, so named because they include rounded mineral fragments called chondrules. Silicate bodies in the solar system are generally assumed to have chondritic ratios of the refractory elements, and until recently chondritic ratios were thought to be the same as solar ratios.

**Clinopyroxene**—A class of minerals that includes diopside and jadeite ( $\text{NaAlSi}_2\text{O}_6$ ) and which is important in eclogite and piclogite.

**Diopside**— $\text{CaMgSi}_2\text{O}_6$ ; an important mineral in basalts and eclogites, but one that is rarer (<10%) in rocks thought to make up most of the shallow mantle.

**Eclogite**—A rock composed primarily of diopside, jadeite and garnet [ $(\text{Ca}, \text{Mg})_3\text{Al}_2\text{Si}_3\text{O}_{12}$ ] and which is the densest common upper-mantle rock. The high-pressure solid form of basalt.

**Enstatite**— $\text{MgSiO}_3$ ; the magnesium-rich component of orthopyroxene [ $(\text{Mg}, \text{Fe})\text{SiO}_3$ ]; a common mineral of the upper mantle.

**Eutectic composition**—The composition of the liquid formed in equilibrium with two or more solid phases at an isobarically invariant point.

**Forsterite**— $\text{Mg}_2\text{SiO}_4$ ; the magnesium-rich component of the mineral olivine.

**Jadeite**—See *clinopyroxene*.

**Kimberlite**—A rare, trace-element-enriched material that erupts rapidly from depths as great as 200 km beneath continental areas, bringing mantle and crustal samples to the surface. Some kimberlites contain diamonds.

**Komatiites**—High-temperature magmas, common in Precambrian times but rare in more recent times, that represent much melting of mantle material. They have

high Mg and Fe contents and often have high Ca to Al ratios.

**Lherzolite**—A rock composed chiefly of olivine but also containing orthopyroxene and generally lesser amounts of clinopyroxene. This rock type is thought to constitute the majority of the upper mantle. It is buoyant relative to garnet-bearing rocks and therefore is concentrated in the shallow mantle. Lherzolites contain little of the basaltic elements such as Ca, Al, Ti, Na and K.

**Majorite**—A cubic, garnet-like form of orthopyroxene, stable in the transition region of the mantle.

**Oceanic plate**—The oceanic crust plus that part of the shallow mantle that, because of their buoyancy and strength, move as a coherent plate in response to the driving forces of plate tectonics.

**Olivine**— $(\text{Mg}, \text{Fe})_2\text{SiO}_4$ ; a greenish mineral that when transparent is the semiprecious stone peridot. Probably the most abundant upper-mantle mineral. The flotation of olivine crystals on dense magma in early Earth history and the sinking of denser garnet-bearing rocks, such as eclogite, have probably enriched the shallow mantle in olivine compared with average mantle compositions.

**Orthopyroxene**—See *enstatite*.

**Peridotites**—A general name for upper-mantle rocks that are composed primarily of olivine and orthopyroxene with subsidiary clinopyroxene and an aluminous phase such as garnet. Lherzolites are a form of peridotite.

**Piclogite**—A hypothetical rock that is mainly clinopyroxene and garnet and which can contain up to 50% olivine by mass. It is intermediate in composition between eclogite and peridotite and similar to komatiite.

**Pyrolite**—A hypothetical rock type formed by mixing about 80% lherzolite and 20% basalt in an attempt to reconstruct "primitive" mantle, prior to melting and melt removal; also a term for undifferentiated mantle. In some models not only is pyrolite the hypothetical original rock of the mantle but most of the mantle is still undifferentiated pyrolite. The alternative point of view is that most or all of the mantle has already been processed to form the crust and the various regions of the mantle.

**Serpentinite**—A rock composed primarily of hydrated olivine [ $(\text{Mg}, \text{Fe})_3\text{Si}_2\text{O}_5(\text{OH})_4$ ], having a silky luster and a soapy feel.

**Transition region**—That part of the mantle between 400 and 650 km depth.

highlands, probably due to the presence of a combination of thick, low-density crustal material and hot upwelling material under each continent. The high surface temperature on Venus makes for a weak lithosphere; a deep convective explanation for the gravity field is required. The continents are probably held up by mantle upwellings or horizontal compression, rather than mainly by the strength of the lithosphere. On Earth, only Africa appears to overlie a deep, hot upwelling mantle current. The other continents have apparently drifted over cold mantle downwellings.

### So where is the crust?

We now know that melting was a common occurrence in the early solar system and, more surprisingly, that the Moon and Mars, at least, have enormous crusts compared with the Earth. The Earth's crust, which is important for

us—it would be impossible to survive on the mantle—is less than 0.5% of the planet. Mars and the Moon have managed to sweat out a large amount of their available crust-forming material. The Moon has been extremely efficient at this. It is thought to have once been covered by a magma ocean in which floated gigantic rockbergs. The Moon then froze, and the magma ocean turned into a solid, hummocky crust covering a thin layer of residual melt that contained high concentrations of many otherwise rare elements, just as evaporation of seawater in inland seas leaves behind a highly concentrated brine. Impacts on the lunar surface excavated some of the Moon's "brine," spreading it over the surface. NASA, in typical fashion, dubbed this brine KREEP, an acronym referring to high concentrations of potassium (hence the K), rare Earth elements and phosphorus. This KREEP, the final liquid dregs of a freezing magma ocean, and the thick crust



(thicker in places than the Earth's crust, on a body only about 1% by mass as large) are the evidence for the Moon's violent, cauldron-like early history. KREEP, although small in total amount, is so ubiquitous in lunar samples and has such high concentrations of otherwise rare elements that it cannot be brushed off as an unimportant oddity. In fact, it forms the backbone of theories of early lunar evolution.

A material similar to KREEP occurs on the Earth. It is equally rare and equally enriched in exotic elements. One could easily overlook this material, kimberlite, except for one important fact: It contains diamonds. Kimberlite is erupted from 200 km deep in the mantle. Because of its rarity it has not played a central role in any generally accepted theory of the Earth. It may be the final liquid dregs of a magma ocean, but if so, where is the equivalent of the thick lunar crust? Also, a magma ocean is the antithesis of current Earth theories, in which the crust and magmas result from heating up and melting of virgin material rather than from cooling down and crystallization of a magma ocean that itself is derivative from the primitive Earth. A magma ocean would imply a chemically stratified Earth, since various crystals would float or sink as they froze out of the melt; the magma ocean itself would have arisen by the separation of melt from more refractory crystals.

So where's the crust? The simplest way out is to assume that what you see is what there is, and that only part of the mantle has been melted and outgassed to provide the atmosphere, ocean and crust. This in fact is the conventional view. The observed concentrations of some gases in the atmosphere and some elements in the crust, however, require most of the mantle to have been processed. Trace element concentrations in some mantle magmas, such as kimberlite, are so high that in tandem with the crust, these magmas contain almost the entire terrestrial inventory of those elements. Although the thick crust expected from a magma ocean is missing, this indirect evidence suggests that a large amount of melting and concentration of trace elements did occur.

Plate-tectonic processes circulate the entire oceanic crust back into the mantle every 100 million years. Erosion also removes part of the continental crust, and some of the eroded material may eventually find its way to deep oceanic trenches, where it is also returned to the mantle. This implies that at any given time only about 10% of the crust is at the surface. Much of the continental crust, however, is more than half the age of the Earth, so one can infer this part has not recirculated recently. The geochemical properties of the oceanic crust do not support the idea of a primitive, unprocessed reservoir or of rapid recycling of subducted oceanic crust back to the surface. The range of isotopic compositions of basalts does not support the idea of a homogeneous mantle. Yet models that call for a primitive and homogeneous mantle are currently in vogue.

I have discussed here a simpler and more satisfactory explanation. Mars and the Moon are small bodies. This means that the pressure in their interiors never gets very high. Layer upon layer of basalt flows can form, building a thick crust, the equivalent of hundreds of kilometers of thickness on the Earth. On the Earth, however, the force of gravity is much higher and the pressure increases rapidly with depth. At low pressures the minerals of the crust are light and the crust-forming melts and minerals rise to the surface of the planet. Melt in the interior eventually makes its way to the crust, some freezing to the bottom of the preexisting crust in the manner of ice forming on a lake, some forcing its way through the crust to form volcanoes, some erupting through fissures to form

rivers and lakes of magma. At moderate pressures, however, equivalent to a depth of 50–60 km on the Earth, but much deeper on Mars, the structure of the low-density crystals changes and they sink rather than float. The base of a thick crust undergoes a similar transformation as it cools. Even if a thick crust formed on the Earth early in its history, its bottom would have dropped off as it cooled and thickened. However, in a deep magma ocean the crustal cations, mainly silicon, calcium and aluminum, along with oxygen, form dense melts and crystallize directly to dense minerals that sink, and only the shallow part of the ocean forms the conventional crust. The final magma "brine" and the more buoyant minerals are trapped in between.

Thus, in my view, the outer part of the Earth has two "crustal" layers. The top layer is normal buoyant crust; the deeper layer is dense crust, denser than normal mantle. The deeper layer probably was originally a dense melt, but it could also be forming continuously by the sinking of eclogite. Olivine-rich rocks, kimberlites and magmas rich in the kimberlite elements come from the center of the sandwich. The lower "crust," however, is dense only when relatively cold. As it warms up, it rises, and the high-density minerals convert to lighter minerals, melt and rise through the overlying mantle, generating new crust at midocean ridges and causing seafloor spreading. To make room for this new crust, the older, colder oceanic crust returns to the mantle, converting to dense crust as it sinks, joining the bottom, dense crustal layer. In conventional views of mantle evolution the subducted oceanic plate either gets stirred back into the mantle or sinks to the core-mantle boundary. In the present scheme the Earth became chemically stratified while it was accreting and this stratification is irreversible.

The lower mantle appears to be more FeO rich and more dense than the upper mantle and the subducting oceanic lithosphere. This prevents surface material from circulating deep into the mantle. Chemical stratification was probably set up initially, during the accretion of the Earth, because of the low density of olivine relative to other products of mantle differentiation. The ideas of an intrinsically dense lower mantle and a chemically distinct transition region have been controversial but are now receiving experimental and theoretical support.<sup>10</sup>

The other solid bodies in the solar system hold clues that are important for understanding the Earth. They record processes in the early solar system and hint at the future of the Earth. The Earth, a middle-aged planet with little memory of its youth, changes slowly. Mars and the Moon are old, dead, unchanging bodies. Venus, our not-quite-identical sister planet, may hold the secrets of how a planet changes with time.

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